Chapter 2

Soil Organic Matter as a Source and a Sink of Plant Nutrients

John M. Duxbury, M. Scott Smith, and John W. Doran
with Carl Jordan, Larry Szott, and Eric Vance

Abstract

Nutrients in soil organic matter (SOM) are present within complex polymers and mineral-organo associations. Investigations into the chemical composition of SOM and the nutrient elements within it have only been partially successful and have not been linked to biological processes. Particle-size and density fractionation studies have provided insight into the architecture of soils and on the distribution and stability of SOM and nutrient elements. Such studies should be continued.

Modern concepts of mineralization and immobilization of N, P, and S in soils recognize that SOM is heterogenous with respect to biological activity, that mineralization-immobilization processes occur simultaneously, and that microbial biomass itself represents a significant sink or source of nutrients. The stability of SOM largely results from protective processes occurring within soils rather than from the creation of recalcitrant chemical structures. Although the stability of SOM is likely to be continuously variable, specific pools are recognized for mathematical modeling purposes. These commonly include microbial biomass (BIO), an unprotected or labile pool (LAB), a pool protected only in the absence of cultivation (POM), and a pool that persists for very long time periods

(COM). At present, the more stable pools are estimated from simulation modeling and cannot be directly measured. Isotopic studies have great potential to help this situation. Questions remain about the validity and utility of both the chloroform fumigation procedures—which are widely used to determine microbial biomass C, N, P, and S—and laboratory N mineralization procedures. In addition, a better understanding is needed of the effects of multiple interacting environmental variables on both microbial dynamics and microbially mediated processes. Most of the effort in understanding the behavior of labile organic matter has been directed to C and N and more attention should be given to factors controlling P and S cycling, which may be partially independent of C and N.

Conversion of land to agricultural use makes a fraction of the protected SOM susceptible to mineralization, disrupts internal recycling of nutrients, and increases the potential for loss of nutrients from the system. Opportunities exist to develop soil and crop management systems that result in better synchrony between nutrient mineralization and plant uptake and conservation of nutrients not used by crop plants. Tillage management, which can be an important tool for the manipulation of nutrient storage and release, has not been investigated to a large extent in the tropical environment.

Soil organic matter is often an important source of negative charge in tropical soils and as such promotes retention of available forms of base cations throughout the soil depth. Many of the negatively charged sites in organic matter, however, are blocked by interactions with Fe, Al, and oxide surfaces. The source of "organic" CEC and its maintenance needs to be identified in terms of the pools of organic matter previously described.

The long-standing issue of whether or not crop yield potential is affected by SOM cannot be resolved solely in terms of nutrient availability. It is important to separate effects due to organic matter per se from those due to its decomposition.

Significant advances in both understanding and managing the behavior of soil organic matter (SOM) as a source or a sink of plant nutrients will only be achieved through studies at a conceptual level. Consequently, this paper considers the mechanisms that regulate nutrient availability rather than empirical or statistical relationships, which abound in both the soil fertility-crop production literature and the ecological literature. Although we recognize the great diversity in tropical soils, we have chosen not to discuss individual soil orders because basic principles apply to all soil types and because we have a poor understanding of how a variety of incompletely characterized biological, chemical, and physical processes interact in many different environments to determine SOM behavior. We have tended to make comparisons between temperate- and tropical-region soils and to discuss agroecosystems because most of the recent advances in our understanding of the composition and behavior of SOM have come from studies in temperate-region agroecosystems. At the same time, we realize that information from temperate-region soils cannot always be extended to tropical soils because differences in soil chemistry and mineralogy strongly influence the behavior of SOM.

It has long been accepted that one of the most important, and certainly the most studied, contributions of SOM to soil fertility is its capacity to supply nutrients for plant growth, especially N. Nutrients are sequestered in, or released from, SOM by two fundamentally different processes: *biological processes* control storage and release of N, P, and S as these elements are contained in structural units of SOM, whereas *chemical processes* control interactions with macro- and micronutrient cations (Ca, Mg, K, Fe, Cu, Zn, and Mn). To some degree these two processes create a paradoxical situation in that mineralization of SOM, which is required to release N, P, and S, reduces the capacity of soils to interact with cations that would be available to plants through chemical equilibration processes. Because SOM is often the major source of negative charge on tropical soils, its maintenance is important for retention of available forms of cations within the soil and must be balanced against any desire to exploit organic N, P, and S reserves.

N, P, and S in Soil Organic Matter

Amounts and Distribution

The accumulation and distribution of SOM and its associated nutrients within the soil primarily depend on the quantity and distribution of organic residue inputs (largely plant residues), on the rates of biological decomposition processes, and on the capacity of the soil to protect SOM (mostly humic substances) from microbial decomposition.

There has been some disagreement about the relationship between latitude and total SOM and soil nutrient storage. Variation in total soil C within the tropics is clearly greater than differences across latitudes. Thus relationships between temperature or latitude and total SOM are difficult to establish unless the influences of mineralogy, management, moisture regime, and other factors are also considered. Sanchez et al. (1982) found no effect of present climate on mean levels of SOM in given soil orders when comparisons were made of agricultural sites. In a study of over 3000 soil profiles from natural ecosystems, Post et al. (1982, 1985) examined total soil C and N as a function of latitude, temperature, and moisture regime. Although their data suggest that soil C and N decrease with decreasing latitude, moisture regime was found to be a more powerful determinant than latitude or temperature. For climatic zones with similar moisture regimes—for example, moist forests—total soil C and N were not consistently correlated with temperature or latitude.

Post et al. (1985) further noted that the C/N ratio was related to temperature, being generally, but not consistently, lower in warmer climates. This results from a more rapid and complete decomposition of labile organic matter. Useful comparisons across climates were possible in this study because all samples came from areas with natural vegetation; no large, comparably homogeneous data base exists for cultivated soils.

Within the tropical environment, the accumulation and distribution of SOM throughout the soil depth is generally affected by ecosystem type. Because inputs of residues to surface soil are higher in tropical forests than in savannas and savannas are more subject to residue loss by fire, more SOM is found in soils under tropical forests than in those under savanna vegetation, with much of the difference being in the top 10 cm of soil.

Generally, 95% or more of the N and S and between 20 and 75% of the P in surface soils are found in SOM. Mean values for the organic C/total N/total S ratio of surface soils are fairly constant across the range of soil types and climates, but generalized differences have been noted between cultivated and virgin soils. Thus, the mean C/N/S ratio of agricultural soils is about 130:10:1.3, whereas that of virgin soils, both grassland and forestland, is about 200:10:1 (Freney, 1986; Stevenson, 1986). These differences can be caused by one or more of the following: (a) preferential mineralization of C relative to N and S, and of N relative to S in cultivated soils, (b) the generally higher nutrient concentration of agricultural crop residues, and (c) differences in retention of the various elements in the soil-plant system after mineralization. The mean value for P is close to that for S, but the P content of SOM is more variable than the other elements. In fact, the C/N, C/P, and N/P ratios can vary widely as a function of parent material, degree of weathering, vegetation, and management, whereas the N/S ratio is less variable and usually within the range of 6-8:1 (Biederbeck, 1978; Freney, 1986). The C/P and N/P ratios of organic matter in tropical soils seem to be especially variable; some are considerably wider (say by a factor of 2) than those in temperate-region soils (e.g., Nye and Bertheux, 1957; Enwezor and Moore, 1966), but others are well within the range found in temperate-region soils (e.g., Bornemisza and Igue, 1967; Islam and Ahmed, 1973; M. J. Jones and Wild, 1975; Neptune et al., 1975; S. J. Smith and Young, 1975; Sharpley and Smith, 1983; Mueller-Harvey et al., 1985). The greater constancy of soil N/S ratios is interpreted to indicate coupling of their cycling in the soil-plant system (Barrow, 1961; Biederbeck, 1978) and the high variability of P to a degree of independence in the cycling of soil organic P relative to C, N, and S. Possible reasons for variability in soil C/P values and for differences in element cycling are discussed later in this chapter.

It has been suggested that highly weathered soils of the tropics (Oxisols and Ultisols) usually contain less total P and have a higher proportion of organic

P than do the young, less-weathered soils of temperate regions. R. A. Olson and Englestad (1972) gave average total P values of 200 mg P kg^{-1} soil for highly weathered tropical soils and 3000 mg P kg^{-1} soil for soils of subhumid temperate regions. The literature supports the 200 mg kg^{-1} figure for tropical soils, but the figure for temperate-region soils seems to be too high; for example, reported values for U.S. soils average about 1100 mg P kg^{-1} (Grove, 1985). Nevertheless, with some exceptions, the trend suggested by Olson and Englestad appears to hold.

Sanchez (1976) considers that 60-80% of the P in tropical soils is usually organic compared to 20-50% for temperate-region soils, but data from the West African savanna (Enwezor and Moore, 1966; M. J. Jones and Wild, 1975; Mueller-Harvey et al., 1985), Brazil (Neptune et al., 1975), and Costa Rica (Bornemisza and Igue, 1967) mostly fall within the range given for temperate-region soils. Whether or not tropical soils contain a greater proportion of their P in organic forms, their reduced total P content and sometimes high phosphate adsorption capacity accentuates the importance of organic P as a source of P for plant growth in these soils.

Forms

Soil chemists have spent a considerable amount of time trying to identify the N-, P-, and S-containing organic structures in soils. The nutrients are almost totally present in complex polymers in associations with other organic and soil mineral components, and isolation of them from whole soil is difficult. The general approach used in these studies has been to subject soil or extracted organic fractions to various separation and chemical procedures, such as hydrolysis, to liberate constituent monomers, which can then be identified individually or by compound class. The present state of this methodology is far from perfect, and considerable loss or alteration of many individual compounds undoubtedly occurs. Research to date is as follows.

Organic N

The work on soil N has been well reviewed by Stevenson (1982b, 1986). In general, only about 40-50% of the organic N in soils can be positively identified as belonging to particular chemical classes. Amino acids and amino sugars are the only quantitatively significant components identified to date. Reported ratios of amino acid N to amino sugar N range from 1 to 46:1, although amino acids usually dominate. Variations in methodology and methodological problems make absolute values and comparisons among different studies highly questionable.

One of the more interesting observations of the chemical work with soil N has been that while newly immobilized ^{15}N has a different fractionation pattern than native soil N (more amino N and hydrolyzable unknown N and less nonhyrolyzable N), it quickly becomes similar to that of the native soil humus. Although interpretation of this observation is subject to the limitations of the fractionation scheme, it suggests that newly added N is rapidly converted to the same chemical structures present in old soil N. If this is correct, the more rapid mineralization of recently added N must be due to differences in substrate availability or protection rather than to differences in structural composition of new and old N. This conclusion is supported by results from long-term field experiments that show little impact of either tillage-induced declines in SOM levels or different cropping systems on the distribution of N forms.

Organic P

Most of the soil organic P is present as esters of orthophosphoric acid, and numerous monoesters and diesters have been isolated from soils (Halstead and McKercher, 1975; G. Anderson, 1980). Small amounts of phosphonates, which contain P bonded to C rather than to O, have been detected in extracts of soil by NMR (nuclear magnetic resonance) (K. R. Tate and Newman, 1982; Hawkes et al., 1984). NMR measurements also show that most of the 30-80% of the organic P that can be extracted from soil is present as monoesters. Phosphate esters of inositols, the cyclohexane analogs of hexose sugars, are the most abundant identifiable compound class, ranging from 5 to 80% of the soil organic P. The hexa- and pentaphosphate forms are the most common in soils. The accumulation of inositol phosphates in a wide variety of soil types is thought to be due to their capacity to form insoluble precipitates with Fe, Al, and Ca and to strongly adsorb on amorphous Fe and Al oxide surfaces (G. Anderson and Arlidge, 1962). In acid soils from Scotland inositol hexaphosphate occupied the same adsorption sites and was more strongly adsorbed than orthophosphate. Inositol-P adsorption was correlated with oxalate-extractable Fe + Al, but adsorption of penta- and hexa-inositol phosphates naturally occurring in soils was best correlated with Fe alone. The quantity of inositol phosphates in Canadian and Scottish soils was correlated with orthophosphate adsorption capacity (McKercher and Anderson, 1968; G. Anderson et al., 1974). A similar result was obtained with 16 acid soils from Bangladesh, where inositol phosphates accounted for 10-82% of the organic P (Islam and Ahmed, 1978). Mineralization of inositol P in two of these soils with high inositol P content was slow unless soil pH was raised by liming or submergence (Islam and Ahmed, 1973).

Other identified components of soil organic P include nucleotides and phospholipids, but both groups are considered to be minor contributors, with nucleotides accounting for less than 5% and phospholipids accounting for less than 15% of the total. Since both groups are components of all living tissues, they may, to some extent, be being extracted from soil organisms and plant roots. Except in soils that contain large amounts of inositol phosphates, structures of phosphate esters in soil are largely unknown.

Organic S

The principal form of sulfur added to soils is amino acid S, mostly composed of methionine, cysteine, and cystine, which are also commonly isolated from acid hydrolyzates of soil and extracted SOM. Sulfur-containing amino acids account for up to 30% of the organic S (Freney, 1986). Between 30 and 70% of the organic S in soils can be reduced to H_2S by HI. Most of the HI-reducible S is thought to be ester sulfate (C-O-S), although C-N-S and C-S-S (as in cystine) bonded sulfur would also be included; carbon-bonded S is not reduced by HI.

Other Structures

From the foregoing summary of our knowledge of the organic forms of N, P, and S in soils, it is evident that at a monomer unit level these elements are found to some extent in structures that are similar, if not identical, to those in which they are added to soil. Exhaustive chemical studies using the best methodology, however, usually identify fewer than half of the specific organic chemical forms of any of these elements in soils. This has led to considerable speculation about the presence of other types of structures, many of which could be formed by chemical reactions involving a wide variety of plant- and microbially derived products. Most of the speculation is over forms of N, e.g., heterocyclic N, including pyridine and phenoxazone derivatives (Flaig, 1975), and while many of these structures would be plausible components of soil humic substances, there is as yet scant evidence that any of them are actually present in SOM.

Comparisons of elemental ratios and structural forms of nutrients present in SOM have not shown any substantial differences between tropical and temperate-region soils. These comparisons do show, however, that such parameters vary widely among soils. Explanations for this variability are lacking.

Organic C, N, S, and P and Aggregate Size

Evaluation of the content and susceptibility to mineralization of organic C, N, S, and P as a function of aggregate size has provided another approach to studying the behavior of SOM. Most studies show that organic C, N, P, and S contents increase and C/nutrient ratios narrow with decreasing aggregate size (Chichester, 1969; Cameron and Posner, 1979; D. W. Anderson et al., 1981; Hinds and Lowe, 1980). In contrast, Elliott (1986) found that the C, N, and P content of aggregates decreased with aggregate size; however, C/N and C/P ratios narrowed with decreasing aggregate size. Some differences in the distribution of the various nutrients have been noted: D. W. Anderson et al. (1981) found that S was preferentially associated with fine clays and Elliott (1986) found greater P accumulation relative to C and N in fine fractions.

Most studies on mineralization of organic matter and nutrients in different sized aggregates (Chichester, 1969; Cameron and Posner, 1979; Lowe and Hinds, 1983) show that both organic matter and nutrient mineralization increase with decreasing aggregate size. However, Elliot (1986) found greater mineralization of C and N in macroaggregates (> 300 μm) than in microaggregates (50-300 μm), and postulated that mineralization of interaggregate organic matter is the main source of nutrient release as SOM levels decline with cultivation. There is also evidence from comparison of cultivated and uncultivated soils that organic matter and nutrients in the fine silt/coarse clay fractions are the most stable. Thus, Tiessen et al. (1983) found that cultivation had the least effect on organic P content in the fine silt coarse clay fraction and Tiessen et al. (1984a) found that the natural ^{15}N abundance of N in this fraction was unchanged upon cultivation. In contrast, organic matter and nutrients associated with fine clays are readily mineralizable. Studies by Ladd et al. (1977a, 1977b) on immobilization and remineralization of ^{15}N in soil also suggest that over time N will accumulate in the coarse clay/fine silt fraction.

Research Needs

In vitro studies of the amounts and forms of nutrients in SOM have little relevance to the behavior of SOM as a source or a sink of nutrients in soils. There is still a pressing need to link chemical and biological studies; chemists tend to fractionate and study components of SOM without regard to the biological significance of the fractions, while biologists and ecologists have developed concepts and models without fully considering the chemical or physical logic of pools of differing biological stability or what can be exerimentally measured.

More information is needed on the chemical forms of organic phosphorus (and perhaps also sulfur) in tropical soils and on their interactions with soil mineral components. Most attention should be given to soils with high phosphate adsorption capacity; these most likely will be acid soils in which strong interactions with active forms of Fe and Al occur. Chemical studies should be extended to determine the bioavailability of organic P and S.

Particle-size and density fractionation procedures provide information at a more detailed level than do studies with whole soil and are beginning to probe the architecture of SOM in soils. Studies along these lines should be continued within a wide range of tropical soils, especially the Oxisols and Ultisols that have exceptional natural soil structure. At the same time, a word of caution: it is important that thought be given to questions of methodology. For example, is air-drying of soils – used by some researchers – appropriate, or does it affect both organic-matter distribution and mineralization? Similar questions should be asked about ultrasonic dispersion.

Biological Mineralization and Immobilization of N, P, and S

Concepts and Models

Gradually over the last few decades, a new and in some ways increasingly complex concept of mineralization and immobilization of nutrients in SOM has evolved. This model attempts to account for, and is in part based upon, the realization that:

1. Microbial biomass constitutes a significant sink and source for nutrients.
2. The process of decomposition is also a process of microbial synthesis.
3. Mineralization and immobilization occur simultaneously in soils.
4. Soil organic matter and the nutrients within it are heterogeneous with regard to biological activity; a fraction cycles very quickly, but some components cycle very slowly.

The process of nutrient mineralization and immobilization represents the net result of several interacting subprocesses, involves many functionally diverse organisms, acts on heterogeneous substrates simultaneously, includes multiple interacting elements, and is subject to multiple interacting environmental controls. This makes it an interesting, worthwhile, and frequent target for simulation models and much of our discussion will be framed in terms common to most of these models.

Environmental Regulation of Mineralization and Immobilization

The important environmental determinants of nutrient turnover and storage via SOM include soil chemistry and mineralogy, soil and vegetation management, and climate.

The effects of climate on biological processes are central to any discussion of the behavior of organic matter in soils. Fairly constant warm temperature is the defining climatic characteristic of tropical soils. More than adequate experimental information exists on the responses of microbial populations and processes to temperature change. Over the temperature range of interest, rates of individual processes increase rather predictably with increasing temperature ($Q10$ = approximately 2). Therefore it is expected that microbially-mediated processes, including turnover of nutrients in plant and animal residues, microbial biomass, and SOM, would generally be accelerated in tropical soils.

Integrating the effects of temperature change on multiple interacting processes and predicting interactions with other environmental parameters in soils is not so simple, however. Attempts to relate overall SOM levels to temperature or to climatic variables in general are a good example. They meet with mixed success because other variables also influence the level of stabilized organic matter in soils, which usually accounts for the bulk of the SOM. Climatic variables are good predictors of residue decomposition processes and of the turnover rate of unprotected organic C, N, P, and S (see later in this chapter).

Effects of Seasonality and Environmental Variation on Processes of Decomposition and Accumulation

It is widely recognized that seasonality and short-term environmental variation are critically important for controlling soil processes. Yet the mechanisms and magnitude of these effects have not been well defined, especially under field conditions. Temperature variation is less in the tropical environment and freeze-thaw effects are unimportant; however, soil wetting and drying cycles may be more extreme in some tropical regions. Fluctuations in both temperature and moisture are known to accelerate organic matter decomposition and N mineralization, although wetting and drying has probably received the most attention (Birch, 1958; Ladd et al., 1977a, 1977b). It is commonly stated that these effects are due to disruption of physically protected organic matter complexes and to accelerated turnover of microbial biomass, but this has not been clearly demonstrated. The short-term accelerating effects of temperature and moisture fluctuations may be quite different from the long-term effects of seasonal variability. It has been suggested that strongly seasonal environments favor accumulation of soil organic matter (for example, Harmsen, 1951). The most plausible explanation is that strong seasonality tends to inhibit decomposition more than primary productivity. Jordan (1985) concludes that as temperature increases, dry seasons will inhibit primary productivity more than decomposition; thus, seasonally dry tropical areas should have less SOM than strongly seasonal temperate zones where subzero temperatures inhibit both producers and decomposers. The data of Post et al. (1982) provide some evidence for this. Some models have considered the effects of environmental fluctuations (van Veen et al., 1984), but it is not certain how hysteretic biological responses should be quantitatively described. This is a difficult but unavoidable problem in describing quantitative relationships between climate and soil processes.

Nutrient Pools and Nutrient Availability

Defining the quality, availability, and activity of organic nutrient substrates and reservoirs is one key to understanding and describing mineralization-immobilization processes. This is a considerably more difficult problem for nutrients in SOM than for nutrients in plant litter. SOM in different soils differs greatly in quality or ability to supply nutrients. It has been apparent, at least since the classic study by Jansson (1958), that SOM cannot be considered homogeneous with regard to potential for mineralization. Isotopic techniques demonstrate that a fraction of the SOM and immobilized nutrients can be remineralized very rapidly, while ^{14}C dating shows that some soil organic C and presumably also the nutrients associated with it have a turnover time of the order of 1000 or more years.

Most current models describe SOM quality and nutrient availability in terms of discrete organic matter fractions, or pools, which vary in activity toward decomposition. Conceptually useful models will include at least pools for microbial biomass, nonliving but available or active C and nutrients, and stabilized or recalcitrant forms of C and nutrients.

Many of the models of SOM dynamics have evolved from the five-pool model of Jenkinson and Rayner (1977), which divided plant residues into decomposable (DPM) and resistant (RPM) fractions and included microbial biomass (BIO) and two forms of stabilized organic matter, named physically protected (POM) and chemically protected (COM) organic matter. Other models (McGill et al., 1981; van Veen and Paul, 1981; Molina et al., 1983; Parton et al., 1983; van Veen et al., 1984; van Fassen and Smilde, 1985) are more complex and use somewhat different terminology, but all have the same general structure. Jenkinson and Raynor's POM pool, and its equivalent in other models is included in models because of the long-term effects of cultivation on SOM levels; i.e., POM is temporarily protected under conditions of no-tillage. For the most stable, oldest COM fraction the mechanism of stabilization has not been clearly defined.

The partitioning of SOM into discrete pools also has some practical mathematical advantages. Complex and difficult kinetics in observed data can usually be adequately described as the sum of multiple components with simple kinetics, generally using first-order reactions. The major limitation to the discrete pool approach is the obvious one, i.e., that the system is likely to be continuously variable. Given this, the concept of pools and the models based on them will never completely match with biological, chemical, or physical reality.

Furthermore, it is not at all clear how these fractions are to be experimentally measured. There can be no completely satisfactory chemical or physical analysis for active nutrients, since availability is conditional, not inherent to the chemical or physical forms of the nutrients – i.e., it depends on interactions with other components and environmental conditions. The size of the active fraction of any nutrient does not define an absolute amount of a specific class of molecules but rather provides a relative index of biological nutrient availability under a particular set of conditions. Experimentally, it seems that this must be approached by biological incubation, perhaps combined with controlled physical manipulation. The most straightforward and widely used measurement of an active nutrient fraction is the Stanford and Smith (1972) aerobic N-mineralization procedure, in which rates of inorganic N accumulation under defined conditions are used to estimate the size of a potentially mineralizable N pool. N mineralization under laboratory conditions was described by a first-order kinetic model, and it was proposed that N-mineralization rates in the field could be calculated by using field temperature and moisture measurements to modify the decomposition rate constant (Stanford et al., 1973; Stanford and Epstein, 1974).

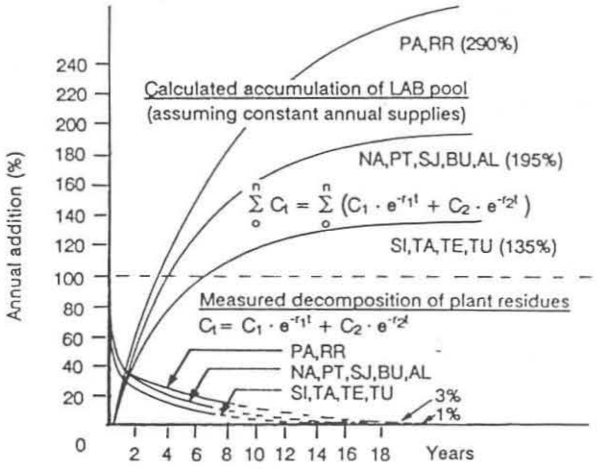
Subsequent studies (Verstraete and Voets, 1976; Cabrera and Kissel, 1988b) and calculations using field temperature and moisture data (J. M. Duxbury, unpublished data) have shown that the method can substantially overpredict N mineralization in the field. Various authors have criticized Stanford and Smith's mathematical methods and choice of a single pool model but have not field tested their proposed alternatives (e.g. Tabatabai and Al-Khafaji, 1980; Naske and Richter, 1981; Talpaz et al., 1981; Juma et al., 1984; Deans et al., 1986). However, Bonde and Lindberg (1988) demonstrated that estimates of N_0 (the soil's N mineralization potential) varied considerably as a function of the regression models employed to describe the N release pattern. Length of incubation (Paustian and Bonde, 1987) and soil pretreatment (Bonde and Lindberg, 1988) are additional factors that affect estimates of both N_0 and rate constants. Beauchamp et al. (1986) and Cabrera and Kissel (1988a) also showed that soil pretreatment, principally air-drying, can lead to higher decomposition rate constants or to a larger potentially mineralizable N pool than in soils that are not dried. The widespread use of air-drying is unfortunate because it likely invalidates the results of a large number of studies and obscures analysis of the value of this approach. For example, possible valid explanations for overprediction of N mineralization include reimmobilization of mineralized N when plants are present in soils, a need for very detailed input of temperature and moisture data, e.g., hourly observations of temperature at small depth increments, and interactions between fluctuating moisture and temperature regimes not included in the procedures for modification of the rate constant.

A more complex but complete measurement of N activity was provided by Paul and Juma (1981). They fractionated organic N into microbial biomass (24-week relative half life), active nonmicrobial biomass (77-week half life), stabilized (27-year half life), and old (600-year half life). Biomass N was estimated directly by chloroform fumigation. Total active N was determined from total N and ^{15}N mineralized during laboratory incubation with the assumption that recently immobilized ^{15}N had uniformly mixed with the active N pool but had entered no other fraction. An old C fraction was estimated from ^{14}C dating and a C/N ratio for this fraction was assumed. Stabilized N was what remained. Although several of the assumptions used in these calculations are questionable, this remains one of the few attempts to assign values to experimentally and conceptually meaningful organic N fractions.

Support for the existence of a young, labile pool of SOM, behaving independently of older SOM pools and not interacting with other soil components, has come from elegant field studies using ^{14}C labeled plant residues (see review by Jenkinson and Ladd, 1981). These studies have shown that the decomposition process can be divided into two phases: an initial rapid phase, which results in loss of about two-thirds of the residue C in 1-2 years in a temperate climate, and a second phase in which decomposition follows a simple first-order kinetic model. The second phase has been followed for 10 years in Germany and for 7 years in Costa Rica with no evidence that first order kinetics will fail to describe the complete decay process (Sauerbeck and Gonzalez, 1977; Gonzalez and Sauerbeck, 1982). The decomposition pattern is not substantially affected by soil properties or crop residue type, at least within the ranges tested (Jenkinson, 1971), and patterns obtained in different climates can essentially be superimposed by using an appropriate rate conversion factor. Compared to results in England and Germany, the decomposition process was four and two times faster at sites in Nigeria and Australia (Jenkinson and Ayanaba, 1977; Ladd et al., 1981).

Knowledge of residue decomposition patterns can be used to predict the size of a pool of labile organic matter (LAB) in soils. The size of the LAB pool depends only on the particular decomposition pattern and the amount of residue inputs (Figure 1). It can be seen from Figure 1 that fairly small differences in the pattern of loss of residue C result in large differences in the size of the LAB pool. The LAB pool will contain partially decomposed plant and microbial residues, live microorganisms, and products of transformation (cf. Paul and Juma, 1981) and presumably contains much of Stanford and Smith's (1972) potentially mineralizable N. The LAB pool probably averages about a quarter to a third of the total SOM in temperate-region soils but would probably be less than this in tropical soils.

Figure 1. Measured decomposition of plant residues and calculated accumulation of a labile (LAB) pool of organic carbon under fallow field conditions for a range of soils and climates in Costa Rica. Figures in parentheses are final equilibrium values. Letters NA, PT, PA, etc., are different soils. (From Gonzalez and Sauerbeck, 1982.)



In most models the physical and chemical significance of the older organic fractions and their function as a source and a sink of nutrients is rather vague. For example, some authors seem to suggest that very old organic matter is recalcitrant because of its chemical nature alone, sometimes referring to this as chemical protection. In other cases the very old material is implied to be protected by chemical interactions with mineral colloids; some people call this chemical protection also, whereas others call it physical protection. In the models considered in this chapter, physical protection is more explicitly considered for the fraction of intermediate age rather than for the very old fraction. The conceptual or experimental basis for this is not apparent.

The chemical structure of organic molecules, by itself, is insufficient to account for the extreme variation in age and turnover times. Although humic molecules are undoubtedly more recalcitrant than natural biopolymers, their intrinsic chemical recalcitrance is much smaller than the observed stabilization of organic matter in soils. Old, humic fractions of soil organic matter, with ages in the thousands of years, have a half life on the order of weeks when extracted materials are added to unextracted soil. It is logical that physical-chemical interactions and protection, like chemical structures, would also be continuously, rather than discretely, variable. We propose that a logical basis for defining pools is the basis of scale. Colloidal or molecular level interactions among organic molecules and minerals, as well as the chemical nature of the organics, are hypothesized to stabilize the most protected, oldest organic components; this fraction could be identified as chemically or colloidally protected organic matter (COM). Structural or intra-aggregate level interactions, for example, organics sequestered in small pores, would lead to intermediate stability; this fraction could be identified as physically or structurally protected organic matter (POM). At least to some extent, the size of these pools would be controlled by different factors: the colloidally protected pool will be regulated by mineralogy and texture, which are invariable over time scales of interest, whereas the structurally protected pool will be controlled by tillage and soil disturbance (Table 1).

This scenario is consistent with the concepts of Tisdall and Oades (1982) and Oades (1984) on the role of different organic constituents in soil aggregation. Roots and fungal hyphae are considered to promote stabilization of macroaggregates, and as such macroaggregation will be controlled to a large extent by soil management practices, such as crop rotation and tillage, which influence the growth and decomposition of plant roots. The stability of microaggregates (less than 250 μm) depends on persistent binding agents, such as complexes of clays, polyvalent metals, and organic matter (which includes polyaccharides and other organic polymers). The stability of microaggregates and their organic binding agents is more characteristically a property of the soil and is less dependent on soil management. Elliott (1986) used this model to demonstrate that N and P in microbial and organic materials, which are

Table 1. Pools of SOM and nutrients, generalized turnover rates, and hypothesized primary controls of pool size

Pool	Turnover time	Pool size controls
Unprotected		
BIO (microbial biomass)	2.5 yr ±, temperate 0.25 yr ±, humid tropics	Substrate availability
LAB (labile)	20 yr ± temperate 5 yr ±, humid tropics	Residue inputs, climate
Protected		
COM (colloidal protection)	1000 yr ±	Soil mineralogy, texture
POM (structural protection)	Depends on physical disturbance	Tillage and aggregate disruption, soil particle-size distribution

normally protected within macroaggregates in undisturbed soils, were mineralized by cultivation of North American grassland soils. Similarly, Skjemstad et al. (1986), using spectroscopic analysis of whole soil and density fractions, concluded that physical protection of SOM, rather than inherent recalcitrance, was the conservation mechanism in undisturbed Vertisols of subtropical Australia.

The nutrient content of the older organic fractions and their contribution to nutrient release is largely undefined except for N, and little information exists on this. Chemical fractionation of SOM gives fractions that vary in age and N content (e.g., Campbell et al., 1967). As the SOM components age they contain less N. If it is assumed that the various SOM fractions are at equilibrium, the rate of mineralization will equal the reciprocal of the age and the relative contributions of the fractions to N mineralization can be estimated on an annual basis. Such a procedure would be quite valuable if a biologically meaningful fractionation scheme could be devised, but it would not be generally applicable to agricultural systems because their SOM components are not at equilibrium.

A new approach to studying the turnover of older SOM components is to use differences in natural ^{13}C abundance levels in SOM arising from differential discrimination between ^{12}C and $^{13}C\text{-}CO_2$ during photosynthesis in C_3 and C_4 plants. This method has been applied to SOM turnover in long-term experiments in which native C_3 plants were replaced by C_4 crop plants, e.g., forest by sugarcane in Brazil (Cerri et al., 1985) or vice versa, e.g., prairie by either wheat or timothy at Sanborn field at the University of Missouri (Balesdent et al., 1988). This approach may provide a means to experimentally test concepts of cycling of old C pools, but it does not directly give information on N, P, or S turnover.

Nutrient Pools in the Organic Matter of Tropical Soils

With regard to nutrient cycling, the effects of climate on SOM quality or lability and on turnover rates of organic nutrients are certainly of greater significance than are the effects of climate on total storage of SOM and nutrients. Few studies directly measure climatic effects on nutrient distribution among SOM pools, yet it is clear that the turnover times of unprotected pools, LAB and BIO, will be accelerated in warm, humid climates but not in dry tropical regions. This is a consequence of more favorable conditions for decomposition and greater C inputs in the humid environment.

Greater rates of soil N mineralization also indicate more rapid nutrient turnover. It has been concluded that net N mineralization, and subsequent nitrification, are greater in tropical forests than in temperate forests (Robertson, 1984; Jordan, 1985; Vitousek and Sanford, 1986; Vitousek and Matson, 1988); however, few tropical sites have been studied and there are large differences among sites, which seem to be related to differences in quantity and quality of litterfall (Vitousek and Matson, 1988). It is not clear what fraction of this increase is derived from mineralization of the large input of plant litter N, as opposed to mineralization of N in SOM.

We propose that the LAB pool is smaller in humid tropical soils than in comparable temperate-region soils. Continuously warm and moist soil conditions would result in more rapid and complete decomposition of unprotected organic materials. However, large and continuous litter input may compensate in part. The tendency for somewhat lower C/N ratios in the tropics, reported by Post et al. (1985), is consistent with more complete degradation of unprotected SOM. One predicted consequence of a smaller LAB pool would be a smaller fraction of the total organic N, P, or S mineralized under any given conditions (for example, in the Stanford and Smith [1972] assay).

The size and turnover of the oldest, colloidally or chemically protected pool (COM) should be relatively insensitive to moisture and temperature but highly sensitive to soil mineralogy and texture. The size of the COM pool will probably be larger in highly weathered, acid tropical soils where mineralogy is dominated by amorphous oxides of Fe and Al – that is, where areas of reactive mineral surfaces are large – than it is in temperate-region soils where mineralogy is dominated by layer silicate clays. However, consistent differences between temperate- and tropical-region soils would not be expected, other than through indirect effects of climate on mineralogy.

Accelerated turnover of unprotected SOM in the humid tropics is of great importance for soil management and soil disturbance. Faster release of nutrients from existing LAB pools could lead to greater nutrient losses if sinks, particularly vegetation, are removed. Also, physical disturbance, and most dramatically intensive tillage, releases a portion of nutrients in the protected POM fraction to the LAB pool. The excellent aggregate structure of soils of the humid tropics suggests that in the undisturbed state POM pools may be at least as large as in temperate soils. Yet once protection is removed, losses of these nutrients should be faster in the humid tropics. In fact, tillage generally results in rapid release of organic N, P, S, and C in the humid tropics (Greenland and Nye, 1959). This explains, at least in part, the common observation that temperate-region Mollisols remain fertile much longer than do tropical-region Oxisols after initial cultivation. This is one sense in which it may be appropriate to speak of humid tropical soils as being more fragile than temperate-region soils.

Microbial Biomass as a Nutrient Source/Sink

The dual role of soil microbes as a catalyst and as a source/sink in nutrient transformations is now widely accepted. Direct measurement of microbial biomass indicates that 1 to 5% of the total organic C and N are stored in living tissue. The importance of the microbial biomass nutrient pool is further magnified by its more rapid turnover than occurs in total soil organic matter. Thus it constitutes a large part of the active fraction. At present, methods for direct measurement of microbial N, P, and S are somewhat less certain than are those for microbial C. The most widely used procedures for all these nutrients are based on the chloroform fumigation-incubation procedure developed for C by Jenkinson and Powlson (1976).

If the rate and extent of substrate degradation are assigned to the catalytic function of microbes, then the source/sink activity can be partitioned into the following factors:

a. The fraction of degraded C that is assimilated into new biomass.
b. The C/nutrient ratios of biomass.
c. Any mass changes associated with biomass maintenance.
d. The rate of cell death or turnover.

We have abundant data from pure culture studies and some soil measurements for the first two factors (a and b), although the extent to which these factors vary in soil and the controls on this variability are poorly defined. The provision for variation in biomass element ratios has been included in some recent models (van Veen et al., 1984), pointing to the sensitivity of mineralization and immobilization to this parameter. Differences in C-assimilation efficiency resulting from tillage systems have also been proposed to be important determinants of soil C storage (Holland and Coleman, 1987). The last two factors (c and d) are difficult if not impossible to investigate directly in soil. Knowledge of the physiology of microbial persistence and viability in natural environments is also very limited.

The current emphasis on microbes as a source and a sink for nutrients should not be taken to indicate that no questions remain to be answered about the catalytic role of microbes. Research in soil enzymology has concentrated on standard determinations of kinetic parameters and temperature response and usually has failed to address the significant questions of how synthesis of these enzymes is regulated, how their persistence is controlled, and how their actual function relates to potential activity (enzyme assays) and variable soil conditions. The energy cost for extracellular enzyme synthesis has not been defined. Microbial uptake systems are very well characterized in pure culture, but their significance is infrequently considered in natural systems. An example where this has been considered and is seen to be highly significant is with regard to uptake of different N sources. It is reasonably well demonstrated that NH_4+ is almost universally preferred over NO_3-, largely removing the latter from N turnover processes. But what is the situation with regard to NH_4+ and organic N sources? Two further questions related to the catalytic role of microbes are considered elsewhere in this chapter: the effects of nutrient interactions on degradation and the mechanisms by which organic substrates are protected from enzymatic attack.

There is no evidence, or reason to suggest, that the size of biomass nutrient pools will be any different in tropical than in temperate-region soils. Qualitative differences in the composition of both the microbial community and soil fauna—that is, differences in species composition—are likely, yet it is doubtful that characterizing such differences is the most fruitful approach to understanding differences in nutrient cycling. It is probable that the turnover rate of biomass nutrient pools will be greatly accelerated in moist tropical climates, particularly where primary productivity is high. There should be much more gross mineralization and immobilization in tropical soils.

These hypotheses are supported by calculations from Paul and Voroney (1983) based on data from cultivated soils in Canada, Brazil, and England (Table 2). Microbial C and N were lowest in Brazil, although not greatly different from the soil in England. Yet C inputs were approximately an order of magnitude higher in Brazil. This and the accelerated rate of decomposition caused estimated turnover times for soil C and microbial biomass to be approximately 10 times greater in the temperate than in the tropical-region soils. N flux through the biomass was thus also about tenfold higher in the Brazilian soil. Although this analysis is obviously limited with regard to number and type of sites, the differences in the controlling parameters, C input and C turnover, will apply to most comparisons of temperate and humid tropical conditions.

The consequences of accelerated biomass turnover and more gross mineralization and immobilization of nutrients would seem to depend on the

Table 2. Amount and turnover rate of C and N for cultivated soils from three locations. (Adapted from Paul and Voroney, 1983.)

	Temperate		Tropical
Determination	England	Canada	Brazil
C inputs ($\overset{Kg}{mg}$ ha^{-1} yr^{-1}) <	1.20	1.60	13.00
Estimated turnover of soil C (yr)[a]	22.00	40.00	2.00
Microbial C (kg ha^{-1})	570.00	1600.00	460.00
Microbial N (kg ha^{-1})	95.00	300.00	84.00
Estimated microbial turnover time (yr)[a]	2.50	6.80	0.24
Estimated N flux through microbial biomass (kg ha^{-1} yr^{-1})	34.00	53.00	350.00

a. Derived from simulation modeling

state of the system. When nutrient source-sink relationships are disturbed – for example, when vegetation is removed – there may be a greater potential for rapid nutrient release and loss. This would not be the case in undisturbed ecosystems or where continuous plant nutrient accumulation is maintained.

It is not clear how accelerated biomass turnover would affect the competition between plants and microbes for nutrients. It could be argued that accelerated gross immobilization would result in greater pressure on plants in terms of nutrient acquisition. Such selection pressure might favor plant-microbial symbioses as mechanisms of nutrient acquisition (mycorrhiza, symbiotic N-fixers). However, net microbial mineralization of nutrients is probably more significant in this regard than are microbial turnover rates. An alternative hypothesis is that the higher and less variable temperatures would make nutrient flux through the microbial biomass fast and constant. This would minimize the significance of microbial biomass as a competitive nutrient sink in tropical soils. Perhaps an indication of this would be provided by the relative nutrient pool sizes of microbial and plant biomass in different climates.

Differences in Cycling among N, P, and S

While the bulk of the N and S in most topsoils, regardless of climate, is held in organic molecules, inorganic forms of P may constitute a large fraction of the total. As soils weather, the significance of organic P as a nutrient source increases. As a generalization, it appears that the significance of P as a soil fertility problem is greater in tropical soils than in temperate-region soils (Vitousek, 1984). Despite the obvious importance of P mineralization and immobilization, much less is known about its behavior than about N.

Sanchez (1976) suggests that a wide C/P ratio is a symptom of P deficiency, but this concept is not universally accepted. Soil organic C/P ratios in tropical soils are highly variable but are not necessarily wider than those of temperate region soils (see preceding discussion). The greater variability of C/P than C/N ratios is often interpreted to mean that P mineralization is uncoupled from C and N mineralization. The much greater susceptibility of soil organic P to alkaline extraction, relative to other elements, also indicates that its storage is somewhat independent of C and N. Since P occurs almost entirely in the ester form (C-O-P) while N is covalently bonded directly to C, it is plausible that their behavior differs. Because S occurs as both C-S and C-O-S, its behavior may involve characteristics of both. McGill and Cole (1981) and Hunt et al. (1983) have used this concept of different bond types to account for interactions among elements and differences in their behavior. It was proposed that the mineralization of esters is regulated by demand for the nutrient, but mineralization of elements covalently bonded to C is regulated by the factors that control use of the energy-yielding substrate.

It is clear that soil organic P can be hydrolyzed extracellularly and mineralized without C degradation, but it may also be assimilated and mineralized with concomitant oxidation of the C. And while the common idea of N mineralization includes microbial uptake followed by oxidation of the C, it is also true that extracellular deaminases exist in soil. In neither case, P or N, has it been conclusively demonstrated what the balance is between extra- and intracellular hydrolysis. More important, the regulation of these processes in soil is poorly documented. For example, how much effect does P, N, or S availability have on mineralization of the same or another element? How variable are the element ratios of microbial biomass and how are these regulated? What is clear is that mineralization-immobilization processes for P and S may differ significantly from those for N and that interactions among these elements need to be considered.

Research Needs

At the heart of any discussion involving differences in climatic conditions is the question of temperature and moisture regulation of soil processes. Describing and understanding these and other stress effects on microbes is in some respects the central problem. Although single-factor response studies (effect of temperature on ..., effect of soil moisture on ...) have been done ad infinitum, we are still not certain how to integrate these to quantitatively predict their effects on mineralization and immobilization of nutrients (or most other processes). This problem will not be resolved simply with multiple-factor experiments (effects of temperature and moisture on ...), but will require greatly improved knowledge of how microbes in soil respond physiologically to various forms of stress.

We propose that all the organic matter in soils is readily mineralizable and that associations within soils, at both the molecular and aggregate levels, prevent this from occurring. Much more attention needs to be given to (a) the chemical nature of associations of organic matter in soils, both between organic molecules themselves and between organic and inorganic soil constituents, (b) the effects of different types of associations on the stability of SOM components to biological decomposition, (c) how soil and cropping management systems influence interactions between organic and inorganic soil constituents, and (d) how the overall architecture of the soil system affects decomposition processes. These studies could be especially fruitful in the tropical soils that have reasonable to high clay contents and exceptionally good physical structure, that is, Oxisols.

Characterizing organic nutrient quality or lability remains a problem, more so in soil organic matter than in plant residues. The chloroform fumigation technique is approaching the status of a universally accepted standard for biomass C measurement. Also, this approach is being widely used for biomass N, and to a lesser extent for P and S. Yet questions remain about its application: What is the most appropriate control? What is the best conversion factor? How do fresh residues and substrates affect the results?

Although the assumptions used in measuring active pools (e.g., Paul and Juma, 1981) can be questioned, such approaches do provide information about the activity of organic nutrients. These tests must be functional, that is, based on biological incubations. It is important to establish widely applicable approaches, if not standard techniques. Although non-tracer techniques, such as the Stanford and Smith (1972) determination of potentially mineralizable N, are of significant value, this area rather than fertilizer efficiency studies is the appropriate place for application of N-isotope studies in tropical soils.

Much less is understood about mineralization and immobilization of P and S than about N. Interactions among elements, such as the effect of P limitation on N storage and cycling, are also poorly described. The significance of P limitation in many tropical soils indicates that these are important problems for future research.

The obvious complexity of mineralization-immobilization makes this an excellent candidate for useful application of mathematical modeling. Yet, a division generally persists between those who model and those who experiment on mineralization and immobilization. There is a need for joint effort between experimenters and modelers to produce models that will both stimulate the development and testing of new hypotheses and which have value at a practical agricultural level.

Management of Soil Organic N, P, and S

Nutrient Cycling in Natural and Agricultural Ecosystems

The soil organic matter content of a given soil in a natural ecosystem approaches a steady state when inputs from plant production balance the decomposition activities of soil microorganisms and soil fauna. The steady state content of C and N attained is governed by the balance between inputs and decomposition rates and the capacity of the soil to protect SOM from decomposition. Although native SOM contents vary widely, they rarely exceed 5-6% except under conditions where microbial decomposition is limited by reduced aeration or where organic matter protection by interactions with minerals is exceptional. For example, with allophane in volcanic soils (Sanchez, 1976; Paul, 1984; Stevenson, 1986). The organic matter content of undisturbed grassland and forest soils changes very slowly; where large changes occur naturally they are usually associated with fire or major climatic changes that affect vegetation. In natural ecosystems, internal recycling of nutrients between primary producers, consumer and decomposer organisms, and abiotic storage pools greatly exceeds flow through the system from atmospheric inputs or leaching losses (Crossley et al., 1984). Agroecosystems differ from natural ecosystems in that energy fluxes, nutrient cycles, and hydrologic characteristics are regulated to varying degrees by physical manipulation of the soil and external inputs of water, nutrients, and energy. The overall impact of these management technologies is disturbance of the normally conservative nutrient cycles and accelerated release of nutrients in SOM to the soil abiotic environment where they can either be taken up by plants or lost through leaching or volatilization. Soil organic matter contents decline until a new steady-state, characteristic of the particular system, is reached. The new steady-state SOM content would be reached fairly quickly in tropical soils if the system were controlled only by the intensity of decomposition processes; however, the rate of release of protected organic matter will also be a major factor because most of the change in SOM levels will be due to changes in the size of this pool.

An important difference in nutrient cycling between natural and agricultural ecosystems is the relative synchrony between plant growth and microbial activity. In natural ecosystems the growth of microorganisms and plants is often in synchrony because they have the same general environmental requirements for growth. Natural selection and species diversity enables biological activity to occur over much of the year, subject only to moisture limitation in the tropical environment. Most available nutrients are used by plants or microorganisms and are recycled through the various SOM pools. Soil contents of available nutrients are minimal during periods of high biological activity (although nutrient fluxes may be high) and nutrients are therefore not susceptible to loss by leaching or volatilization during these times. In agricultural systems, high crop demand for nutrients occurs during defined time intervals, and there may be considerable periods of time when plants are absent. If release of nutrients is insufficient to meet crop demand, supplementary fertilizer may be needed. In the same system, nutrients mineralized during periods when crops are not growing will accumulate in soil and, together with residual fertilizer, are subject to loss by leaching or volatilization. Even with continuous cropping the root systems of annual crop plants require some time to develop and are never as extensive as those of perennial plants; hence, annual crop plants are unlikely to be as effective at recovering nutrients from soils as perennial crop plants or natural perennial vegetation. Consequently, many agricultural ecosystems are inherently more nutrient-leaky than are natural ecosystems. More attention should be given to including the natural recycling capacity of perennial plants in agricultural systems. Agroforestry systems are an example of where this issue has been addressed.

One approach (McGill and Myers, 1987) to predict synchrony between N mineralization and crop uptake of N is shown in Figure 2. Here N mineralization rate is assumed to depend only on soil moisture and temperature regimes, which are combined into a single index. On this basis synchrony is predicted to be reasonable for crop growth in temperate North America but N mineralization both precedes and continues beyond crop growth. In subtropical and tropical environments, where more than one crop may be grown, predicted synchrony varies considerably, suggesting that there may be a need to add fertilizer N at some times and to conserve mineralized N at other times. This analysis does not include factors that alter the availability of protected SOM pools, but it identifies a probable need to develop nutrient conservation strategies. Evaluations of this kind illustrate the potential utility of even simple models and also the need to determine actual N, P, and S mineralization rates under field conditions.

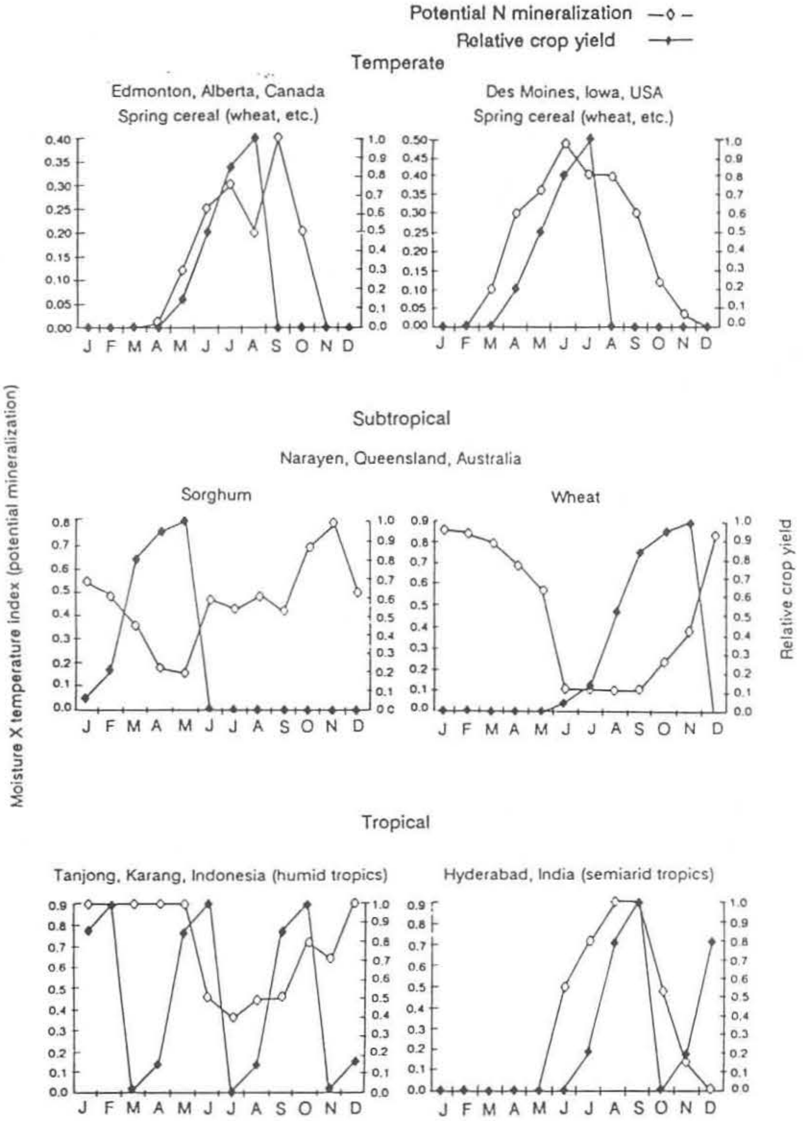
Management of Nutrient Storage and Release from Different SOM Pools

The Labile Organic Pool

As a first approximation, the size of the LAB pool will be controlled by the quantity of residue inputs and climate. Several reasonably long-term agronomic field experiments demonstrate that, as predicted, the size of the LAB pool is directly proportional to residue inputs. C retention in an 11-year study in Iowa (Larson et al., 1972) was 21% of that added at various residue addition rates up to 16 Mg ha^{-1} yr^{-1} with no effect of crop residue type (among alfalfa, corn stalks, oat straw, and bromegrass) on C retention. In a 20-year experiment at Sameru, Nigeria, M. J. Jones (1971) found C retention of 14% and N retention of 30% for additions of farmyard manure up to 12.5 Mg ha^{-1} yr^{-1}.

More research may well show that other variables will affect residue decomposition patterns and hence the size of the LAB pool, but the general principle that the size of this pool is proportional to the amount of residue added

Figure 2. Effect of climate and cropping on synchrony between relative crop yield and potential for mineralization of soil N based on a soil moisture x temperature index. (Adapted from McGill and Myers, 1987.)



will hold. Factors likely to affect the decomposition pattern and C retention include the physical state of the residue, where it is placed in soil, and the method and frequency of tillage. Residue chemical composition and soil variables that affect the activity of decomposer organisms may also be important. Animal manures are 3 to 4 times more effective than plant residues at increasing the LAB pool because they have already undergone the rapid initial decomposition phase.

In absolute terms, the size of the LAB pool derived from plant residue additions will be smaller in tropical soils than in temperate-region soils, where a value of about 2.5 times the annual residue addition rate has been established (Sauerbeck and Gonzalez, 1977). Using this as a maximum value, and with constant annual residue inputs of 1000 kg C ha^{-1}, the LAB pool would contain a maximum of 2500 kg C ha^{-1}, or 250 kg N ha^{-1} and 25 kg P ha^{-1} if the C/N and C/P ratios are 10:1 and 100:1, respectively. It also responds quickly to changes in residue inputs so that current agricultural management is important.

One question on which there is little information is whether the nutrient content of the LAB pool is influenced by the amounts of nutrients in added residues or fertilizers. When various crop residues containing approximately equal amounts of C but different amounts of N, P, and S were incorporated annually into a soil cropped to maize, soil organic C, but not N or S levels, were increased by essentially identical amounts after 11 years (Larson et al., 1972). Soil organic N and S levels appeared to increase with increasing content in the residue, although incomplete residue analysis and fertilizer addition complicate this interpretation. Residue addition had a relatively small effect on soil organic P content; for example, the addition of 16 Mg of alfalfa ha^{-1} yr^{-1} caused soil organic C to increase by 48%, organic N by 41%, and organic S by 49%, whereas soil organic P only increased by 14%. Besides reinforcing the concept that soil organic P behaves somewhat independently from C, N, and S, this experiment strongly suggests that the nutrient content of crop residues will affect the nutrient content of the LAB pool. Fertilizer additions may have a similar effect, but we are not aware of data that would demonstrate this for N or S; for P, however, long-term additions of fertilizer P (35 kg P ha^{-1} yr^{-1} for 125 years) increased the organic P content of SOM in the Park Grass experiment at Rothamsted, England, by 1.5 times without altering SOM contents (Hawkes et al., 1984).

Protected Organic Matter Pools

The size of the LAB pool of SOM and the controls over this pool are such that major changes in SOM contents cannot be achieved by manipulation of this pool alone. In fact, the large declines in SOM associated with tillage of soils and the gains in organic matter when tillage is stopped are mostly associated with the POM pool although turnover of the LAB pool may also be affected. The most effective way to increase SOM in cultivated soils is to stop tillage. This, of course, immediately reduces the availability of nutrients to crops but results in gradual accumulation of nutrient reserves in SOM. Shifting cultivation in the tropics operates on alternating phases of exploitation and accumulation. In effect, similar systems are widely practiced in the temperate regions on a compressed time scale through the use of short-term crop rotations. The most effective rotations are those in which a forage legume is grown to accumulate N via biological N fixation in the no-till phase of the rotation. This system leads to rapid accumulation of N and hence to a short-term rotation.

Although tillage is in many ways the most powerful tool available for the manipulation of storage and release of nutrients in SOM, its effects on the soil system are complex. Tillage and crop residue management practices have a large influence on soil temperature, water, and aeration regimes and on the spatial and temporal availability of energy and nutrients to microorganisms. The redistribution of organic matter and soil organisms with reduced tillage is a major factor responsible for greater recycling and retention of N than with conventional tillage (House et al., 1984; Fox and Bandel, 1986). Organic matter contents of the upper layers of soil, microbial biomass, and reserves of potentially mineralizable N are often significantly higher with no-tillage than with moldboard plow tillage (Table 3). Such increases in microbial biomass and activity and organic N reserves result from surface placement of residues, accumulation of SOM, and a more optimal water status for biological activity in the surface 0-10 cm of reduced-tillage soils in both temperate and tropical regions (Ayanaba et al., 1976; Doran, 1987a). Increased microbial biomass is also associated with increases in plant root activity near the surface of no-tillage soils (Lynch and Panting, 1980; Carter and Rennie, 1987).

Table 3. Effects of tillage on soil water content, chemical components, and soil microbial biomass as a function of soil depth at six (four continuous maize, two wheat/fallow) long-term (6-13 year) experiments in the United States. (Data adapted from Doran, 1987a)

Soil parameter	Ratio of values for no tillage to moldboard plow tillage			
	0-7.5 cm	7.5-15 cm	15-30 cm	0-30 cm
Water content	1.28*	1.08	1.08	1.13
Water-soluble C	1.47*	0.98	1.24	1.23
Total organic carbon	1.42*	1.00	0.94*	1.06
Total Kjeldahl N	1.29*	1.01	0.97	1.06
Potentially mineralizable N	1.37*	0.98	0.93*	1.05
Microbial biomass	1.54*	0.98	1.00	1.13

* Significant difference ($p < 0.05$) using F-test for tillage across locations

The magnitude of management-related changes in the properties of tilled layers and microbial responses can greatly depend on previous management, cropping, and degree of tillage. The levels of organic C, N, and microbial biomass in surface soil of a winter wheat (*Triticum aestivum* L.)/fallow rotation in western Nebraska were inversely related to degree of soil tillage during fallow (Doran, 1987b). In previously cultivated land where initial SOM levels were lower, these differences were much less pronounced than where tillage comparisons were initiated in native grass sod. Also, over an 11-year period, the total soil N content with no-tillage management was 9% greater than when crop/fallow was first initiated. In converting from grassland to wheat/fallow, declines in soil organic C and N levels, regardless of tillage management, reflect decreased inputs of C and N resulting from reduced plant production and residue inputs. In either case, however, reduced tillage has conserved more N in the upper layers of soil—through reducing net mineralization of crop residues and SOM—than has subtillage or plowing.

Tillage-induced differences of N in the upper layers of soil reserves of potentially mineralizable N and microbial biomass may vary with climate and cropping management practices. Differences in mineralizable N reserves between plow and no-tillage management at six long-term experiments across the United States ranged from 12 to 122 kg N ha^{-1} and were highly correlated with mean annual precipitation (Doran, 1987b). These trends likely result from increased cropping intensity and plant productivity associated with increasing rainfall. Differences between tillage management were least and values for potentially mineralizable N lowest for wheat/fallow production in a low-rainfall region. Higher contents of mineralizable N and greater differences were observed at four locations under continuous cropping with maize, especially at the most humid location where a rye cover crop was also planted.

Interactions between microbial activity and mineralization of soil organic N are often controlled by environmental factors. Predicting how changed environmental conditions in reduced tillage soils will affect net mineralization is difficult because the contrasting effects of increased water and reduced temperatures on net mineralization may vary during the growing season and across climates (Fox and Bandel, 1986; Doran and Smith, 1987). Also, increased microbial biomass levels in no-tillage surface soils during the growing season can serve as a sink for immobilization of N. Greater immobilization of fertilizer N has been observed in no-tillage crop production than in production using plowing or shallow tillage (Rice et al., 1986; Carter and Rennie, 1987).

The effectiveness of tillage in releasing the N contained in soil microbial biomass and organic N reserves is also influenced by soil type and the density of plant roots. The productivity of grass pastures in subtropical Australia is often limited by reduced availability of mineral N as a result of accumulation of root and plant debris with a high C/N ratio and increased immobilization of N in microbial biomass. Periodic cultivation of grass pastures increases mineralization of soil N and stimulates grass production through changes in root density and mineralization of microbial and organic N reserves (Table 4). In clay soils the N mineralized by cultivation may come largely from stabilized forms of organic N, whereas in sandy, loamy, coarse-textured soils microbial biomass may be the predominant source of mineralized N. Changes in microbial biomass resulting from cultivation paralleling those for root biomass suggest an association between changes in root density and microbial biomass levels in soil.

Table 4. Effect of soil cultivation on N budgets for the 0-30 cm soil depth interval of grass pastures at two sites in Queensland, Australia. (After Doran, 1987b.)

Plant or Soil Component	Green panic, clay (kg N ha^{-1})			Buffelgrass, sandy clay loam (kg N ha^{-1})		
	No tillage	Chisel plow	difference	No tillage	Plow/ resown	difference
Plant nitrogen						
Tops	45	62	+17	28	41	+13
Roots	84	102	+18	116	94	-22
Soil nitrogen						
Nitrate + ammonium	4	10	+6	8	7	-1
Mineralizable N	945	882	-63	480	459	-21
Microbial biomass N	318	332	+14	153	116	-37
Total organic N		8143			3144	

Research Needs

Agricultural management systems control residue placement, soil disturbance, and the *in situ* production of soil organic matter. The resulting effects on soil physical properties and substrate availability greatly influence plant and microbial activities and the cycling of C, N, and other nutrients. Interactions between the soil physical, chemical, and biological characteristics with various systems of organic matter management are greatly influenced by climate, soil, and initial SOM levels. Development of appropriate management strategies for most efficient utilization of N and other elements requires better understanding of these interactions, especially in tropical ecosystems where little research of this nature has been previously conducted. Suggested research needs are to:

1. Determine the short- and long-term effects of tillage and crop management systems on organic nutrient reserves in tropical soils. In particular, what role do surface placement and lack of tillage play in the accumulation of occluded organic matter (POM) and nutrient reserves? Will immobilization of nutrients in POM decrease with time in undisturbed soils, resulting in more recycling and less nutrient tie up?
2. Identify the processes responsible for the tillage-induced declines in SOM. The continual decline over many years, coupled with several observations that sieved virgin soils and soil cores mineralize N at the same rate, suggests that soil disruption by the tillage instrument is not the primary driving force. Knowledge of the mechanisms involved in the tillage effect might lead to more controlled nutrient-release practices, may aid the design of tillage tools, and may provide a basis for choosing between alternative tillage methods. A possible cause for the slowness of the tillage effect is the gradual exposure of new soil to the more highly physically disruptive environment at the soil surface. The central issue may prove to be one of scale, i.e., tillage equipment operates at a macroscale whereas the forces generated, for example, by drying and rewetting act at a microscale level.
3. Better define the function of soil organic matter. Both biological and physical-chemical functions, need to be considered. What is the optimum balance between exploitation of SOM as a source of nutrients and energy and its conservation as a conditioner of the soil physical environment? More emphasis on comparisons of virgin and cultivated soils may provide insight into mechanisms of changes resulting from agricultural management.
4. Include several elements in most studies on SOM. Much of the work on organic matter dynamics, for example has been limited to C and application to N, P, and S is by inference. Almost all the work on the effects of tillage on nutrient release and storage has dealt only with N. Future studies should include several elements to establish similarities, differences, and interactions among and between them. Such information is likely to influence decisions on management of SOM.
5. Better define the concept of a biologically inert pool of organic matter, i.e., the COM pool. This concept is intuitively unsatisfactory. What could be the mechanisms that so completely protect inherently decomposable organic molecules from decomposition? Surely there can be no one mechanism because soil inorganic components vary so greatly as a function of weathering; or, are overriding interactions with Al and Fe species common to all soils?

Organic Matter and Cationic Nutrients

Mechanisms of Interaction

Charge development on SOM is pH dependent and is predominately negative. The principal functional groups involved in negative charge development are carboxylic and phenolic acids. Some positive charge development through protonation of amino groups can also occur, but it is quantitatively small compared to negative charge development. Interaction of cations with SOM involves several different bonding mechanisms. Basic nutrient cations (Ca, Mg, and K) interact predominantly via electrostatic attraction, whereas divalent transition metal nutrients (Zn, Mn, Fe, and Cu) and trivalent Fe form coordinate linkages of varying strength with SOM. Nonnutrient cations, especially Al, will compete for coordination sites. Organic matter-metal complexes may be monodentate, bidentate, and so on, and they may include the formation of cyclic chelate structures. Studies of cation binding to SOM are usually carried out on organic matter extracted from soil and fractionated into the traditional fulvic and humic substances. The strength of binding of various metals to soil humic substances usually follows the order Fe (III) > Al > Cu > Ni > Co > Pb > Zn > Mn > Ca > Mg, but the order somewhat depends on pH. The strength of binding of a given metal also depends on the amount added, indicating that a range of organic complexing sites, with different affinities for the metal, are involved.

Importance of Organic Matter/Cation Interactions in Tropical Soils

Cation Exchange and Base Retention

The CEC of soils is especially important for retention of the basic cations (Ca, Mg, and K) that interact with soil surfaces via electrostatic attraction. Both soil minerals and SOM contribute to CEC in soils; however, as soils are weathered, their CEC drops due to changes in mineralogy from 2:1 type layer aluminosilicate minerals to kaolinite and amorphous oxides of Fe and Al. The CEC of most soils dominated by 2:1 layer minerals is usually within the range of 15-30 cmol_c kg^{-1} soil, whereas that of soils dominated by kaolinite and amorphous oxides is almost always less than 5 $\text{cmol}_c\ \text{kg}^{-1}$ soil and most of that may be associated with SOM rather than with the mineral components. Because there are large areas of highly weathered soils in the tropics, maintenance of SOM to provide CEC is more important in tropical than in temperate regions.

Approximate average values for the total acidity of extracted soil humic substances are between 700 and 1000 $\text{cmol}_c\ \text{kg}^{-1}$ organic matter (Schnitzer, 1978), indicating a high potential for SOM to add CEC to soils. However, much of this charge is not expressed in soils, both because it is pH dependent and because many of the negatively charged sites are blocked by interactions with

Al and perhaps also Fe. The amount of CEC contributed to soil by SOM is best measured by the multiple-regression technique used by Helling et al. (1964) since this avoids extraction of organic matter from soil. In soils from Wisconsin, SOM contributed about 180 $\mathrm{cmol_c\ kg^{-1}}$ organic matter at pH 5 and changing pH altered CEC by about 30 $\mathrm{cmol_c\ kg^{-1}}$ organic matter per pH unit. The maximum charge development of about 350 $\mathrm{cmol_c\ kg^{-1}}$ organic matter (at pH 8.2) was considerably below that expected from *in vitro* measurements with soil humic substances and indicates the importance of nonexchangeable interactions with soil Al and Fe. Similar methodology has not been applied to highly weathered soils, but it is likely that the CEC of organic matter in such soils will be lower because a greater degree of blockage of negatively charged sites by Al and Fe can be anticipated. Relationships between CEC and SOM content for highly weathered soils of the Brazilian Cerrados were found to depend on soil pH (Lopes and Cox, 1977). SOM was only effective at increasing CEC levels above a pH of 5.5, which is consistent with blockage of exchange sites by either Al or Fe at lower pH values. This result suggests that attempts to increase the CEC of acid soils by increasing SOM levels may meet with short-term success but long-term disappointment as the soil system equilibrates to block newly formed exchange sites. Perhaps the approach that should be taken would be to raise soil pH by liming, but it is not clear that this will unblock previously blocked sites.

It is often considered that one benefit of the older, more biologically inert pools of SOM is that they provide CEC to soils; however, it is possible that these pools are also fairly inert on a chemical basis and contribute little to soil CEC. If the LAB organic pool were to be the most important organic source of CEC in tropical soils, maintenance of this pool at high levels would be an important objective for maintaining CEC.

Despite an incomplete understanding of the relationships between SOM and soil CEC and how to manipulate these relationships in acid soils, it is clear that destruction of SOM will reduce CEC levels and have a negative impact on base cation retention.

Micronutrient Availability

Organically complexed micronutrient metals, and perhaps also borate complexes with soil carbohydrates, are generally considered an important component of the labile reservoir of these elements in soils. In addition, soluble organo-metal complexes are often a major proportion of the micronutrients in soil solution and aid in the transport of micronutrients to plant roots. This is most important in high pH soils because, with the exception of B and Mo, the solubility of inorganic forms of all of the micronutrients decreases with increasing pH. Micronutrient deficiencies are common in tropical regions that have high pH soils which have been cultivated for long time periods, such as India, and this situation would likely be improved by increasing SOM contents.

Indirect Effects on Nutrient Availability and Soil Fertility

Complexing of organic matter with reactive Al and Fe surfaces reduces soil CEC but has beneficial effects in that it blocks these sites and reduces the capacity of soil to fix phosphate and sulfate. Increasing SOM may also stimulate desorption of phosphate and sulfate by acting as a competing anion. However, this may not be very useful if the inorganic anions are simply replaced by organic esters of these same elements. An unequivocal effect of complexing of Al by SOM is that it reduces exchangeable Al levels, which is important because Al toxicity is a common problem in acid Oxisols and Ultisols. Strong complexing with Al and Fe is probably also the reason that Oxisols and Ultisols have such good physical structure.

Overall, there are good and bad sides to the complexing of SOM with soil inorganic constituents but reduction in SOM contents has negative effects on all of the soil properties discussed in this section.

Research Needs

The case for developing a better understanding of SOM interactions with soil mineral components has already been stated in terms of their effects on nutrient release by mineralization processes. It is also important that such studies include an analysis of how interactions involving the various organic pools affect the contribution of these pools to cation retention and availability in soils.

Organic Sources of Nutrients and Crop Yield

The importance of SOM and organic residues to crop production has diminished in developed countries over the last 40 or so years with the availability of cheap energy. In these countries, chemical fertilizers are the main source of the major plant nutrients, and other practices that help to overcome adverse effects associated with lowered SOM contents have been adopted—for example—drainage and irrigation to correct poorer soil physical conditions. High-energy input agricultural systems are not, however, economically viable in many of the developing nations of the tropics where agriculture relies more on SOM, animal manures, and crop residues to supply needed nutrients.

Chemical fertilizers, in fact, are generally needed to supply sufficient N, P, K, and sometimes other nutrients to reach the yield potential of improved crop varieties. The slow release of nutrients from SOM is often considered advantageous (Sanchez, 1976; Avnimelech, 1986) for both soluble nutrients, such as N, and for nutrients that may interact with soil surfaces and become less available, such as P. It would seem that one advantage of chemical fertilizers over organic nutrient sources, however, is that nutrient forms, amounts, placement, and timing of addition are easily regulated and can be matched to the needs of a particular crop. Given best management practices, one would expect greater crop recovery of relatively mobile nutrients, such as N, with chemical fertilizers than with organic nutrient sources and this is generally so. It is less clear whether nutrients that chemically interact with soil surfaces, such as P, are best supplied by organic nutrient sources or by inorganic fertilizers.

A long-standing question is whether or not crop yield potential is affected by SOM level. Although most agronomists would consider that this is not the case for the main grain crops of the world, there are several recent studies that show higher yield potentials with organic manures than with inorganic fertilizers. For example, farmyard manure gave higher yields of both small grain and root crops than were achieved with fertilizers in the long-term experiments at Rothamsted and Woburn, England (Johnston and Mattingly, 1976; Cooke, 1977) and of maize in Colorado (S. R. Olsen, 1986). Similar results have sometimes been reported for paddy rice with various organic nutrient sources, including farmyard manures, green manures, and composts (Chatterjee et al., 1979; Tiwari et al., 1980; Guar, 1984; Kumazawa, 1984; Oh, 1984; Bhatti et al., 1985; Bouldin, 1988). Positive effects of organic nutrient sources seem to be most often observed with crop varieties that have high yield potential. Several of the authors cited also report better efficiency of nutrient use when organic nutrient sources are used in combination with inorganic fertilizers. The focus of most studies has been on N supply, and it has been suggested that the beneficial effects of the organic nutrient sources are due to their increasing the availability of P, K, or micronutrients (Avnimelech, 1986; Bouldin, 1988) or to better root development (Kumazawa, 1984). The release of N from organic sources late in the growing season is also considered important for high yields of rice, and S. R. Olsen (1986) believes that the supply of ammonium from decomposing organic matter provides a balance of ammonium to nitrate which results in better crop growth.

While there is ample evidence to suggest that organic amendments to soil can improve crop yield, this result is not routinely found; in fact, it is well known that crop yield can be depressed if there is net immobilization of nutrients during microbial decomposition of organic residues. Management of organic manures in flooded soils is not straightforward, and loss of nutrients from both the soil and the residue and/or generation of low molecular weight organic acids toxic to plants can occur. In some of the studies in which organic nutrient sources proved to be superior to inorganic fertilizers, it was suggested that the effect was due only to better supply of nutrients from the organic matter. Such a claim is extravagant as it is impossible to monitor all the effects of the organic matter on nutrient availability let alone on the soil-plant system as a whole. Besides effects on nutrient supply, which can be direct (nutrients supplied by the added organic matter) or indirect (effect of the added organic matter on the availability of soil nutrients), the organic matter will presumably also have varying effects upon soil physical properties and microbial and animal life in the soil. The experience with organic manures can to some degree be extrapolated to SOM, but with so many interacting factors in the soil-plant system it is clearly difficult to conclusively establish cause-and-effect relationships between either organic manures or SOM, nutrient supply, and crop yield. As noted out by Bouldin (1988), it is important to separate effects due to organic matter per se from those associated with its decomposition.

Research Needs

While decomposition of SOM and organic manures promotes plant growth through release of nutrients, microbial activity will have additional, less easily definable, impacts on the soil-plant system. Assessment of these additional impacts is essential not only to interpretation of experiments in terms of nutrient availability but also to attaining an overall understanding of the behavior of soil-plant systems. Without this understanding, management choices will be made on an empirical rather than a mechanistic basis and will have reduced success.

Conclusions and Recommendations

Soil organic matter impacts nutrient availability via active (biological transformation and mineralization) and passive (chemical) processes. Our knowledge of the dynamics of SOM and its associated nutrients has developed considerably in recent years. The concepts of pools of organic matter differing in susceptibility to biological decomposition, each responding differentially to soil, environmental, and agricultural management variables, has provided the foundation for understanding how SOM functions as a source and a sink of plant nutrients and how this function can be managed. Nevertheless, while concepts and models are in place, we have a limited capacity to directly measure the various organic matter pools and a limited understanding of the linkages between pools, In addition, we lack experimental verification for proposed mechanisms of protection of organic matter in soils.

Although the focus of future research should be on the dynamic aspects of nutrient availability, we must not forget the passive functions of SOM, which should also be evaluated in terms of the "pool" concept. For example, if the contribution of organic matter to soil CEC is largely associated with the old, stable organic matter pool, there is probably little opportunity for management of this property. If, however, the labile organic pool contributes significantly to CEC, maintenance of this pool, which is strongly affected by recent management practices, would be an important goal.

Differences in soil composition arise as a function of weathering of soils and these, coupled with climate, have a marked effect on the content and behavior of organic matter in soils. Consequently, in a general sense, there are differences in the behavior of SOM in tropical and temperate-region soils because of differences in weathering of soils and climate. At the same time, the biological, chemical, and physical principles governing the behavior of SOM are the same for all soils, and seemingly disparate soils may show remarkable similarities in a particular property; an example of this is the relationship between soil inositol phosphate content and phosphate adsorption capacity.

Because the scope of this chapter is rather wide, we have identified specific research needs at the end of each section. More general recommendations for future research follow.

Theme Imperative: Define the processes controlling nutrient release and storage in SOM and their regulation by soil, climatic, and management variables.

Research Imperatives:

1. Undertake comparative studies of nutrient cycling at sites carefully selected for variation in temperature and moisture regimes, soil mineralogy, native vegetation (natural ecosystems), and agricultural management.
2. Initiate long-term (decade scale), *in situ* isotopic tracer studies, using multiple labels to probe C and nutrient dynamics of labile pools and interactions between labile and more stable organic matter pools.
3. Because of the significance of phosphorus limitations to plant growth in the humid tropics, give high priority to determining the forms, interactions, and biological transformations of soil organic phosphorus.
4. Develop better ways of describing the quality or lability of nutrients in soil organic matter. Ensure that conceptual pools are meaningful in terms of biological activity and are experimentally measurable. Give more attention to extending laboratory-measured properties, such as N mineralization, to *in situ* nutrient behavior.
5. Define mechanisms of protection of soil organic matter and nutrients from mineralization and quantitatively relate them to soil chemical and physical properties and environmental variables.
6. Evaluate the biotic and abiotic regulation of microbial assimilation and release of nutrients and their relationships to soil nutrient dynamics.